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AN ESTIMATE OF THE MINIMUM POSSIBLE SURFACE TEMPERATURE AT THE SOUTH POLE

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ABSTRACT

Computations are made of the surface and tropospheric cooling which could occur during the 6-month winter night at the South Pole if only radiation-exchange processes were operating. Actual soundings taken at the beginning and end of the 1957 winter season indicate that this assumption is hardly realistic for the troposphere, but may be more applicable to the ozonosphere.

1. INTRODUCTION

The subject of extremes of temperature and other climatic elements which have been or could be experienced on the surface of the earth is one which holds an interest for most people—both scientists and laymen alike. The question of low temperatures which may occur at the South Pole during the long polar night has no doubt been the subject of much speculation. However, the establishment of a scientific station at the South Pole by the U. S. National Committee for the International Geophysical Year in January 1957, has given substance to the question. From the continuous records which are being taken, we shall at last have a quantitative measure of the annual temperature cycle, and learn the extremes to which the thermometer will dip, at least during the winter seasons of 1957 and 1958. There will, of course, always remain the question of how cold *can* it get at the South Pole.

Occurrence of an extreme low temperature near the ground at the South Pole (or at any other location) necessitates the simultaneous occurrence of an optimum combination of several meteorological elements; absence of solar radiation, clear skies, and calm air are the most essential requirements, with the ultimate fall in temperature dependent upon the duration of these conditions.

In the work to be described, virtually optimum circumstances have been assumed to persist during the polar night (about 180 days), and the resulting "minimum possible" surface and tropospheric temperatures have been determined. With such background information, it is hoped that when the complete data from the IGY South Pole Station have been obtained, it may be possible to make a better quantitative analysis of the effects of the various and particular meteorological influences governing the observed wintertime temperature regimes, than would otherwise be the case.

2. THE MODEL

The computations and results which are to be described were formulated and are to be interpreted under the assumption that the snow-covered ground and the troposphere constitute a partially closed system. No external sources of energy were provided, such as would be represented by advection of warm air into the region, or a flux of sensible heat to the surface from the underlying snow. However, the upper portion of the troposphere was assumed to represent a region of heat loss in the sense that there was no downward flux of radiation into that layer which might compensate, at least in part, for the loss of radiation to outer space from the lower layers.

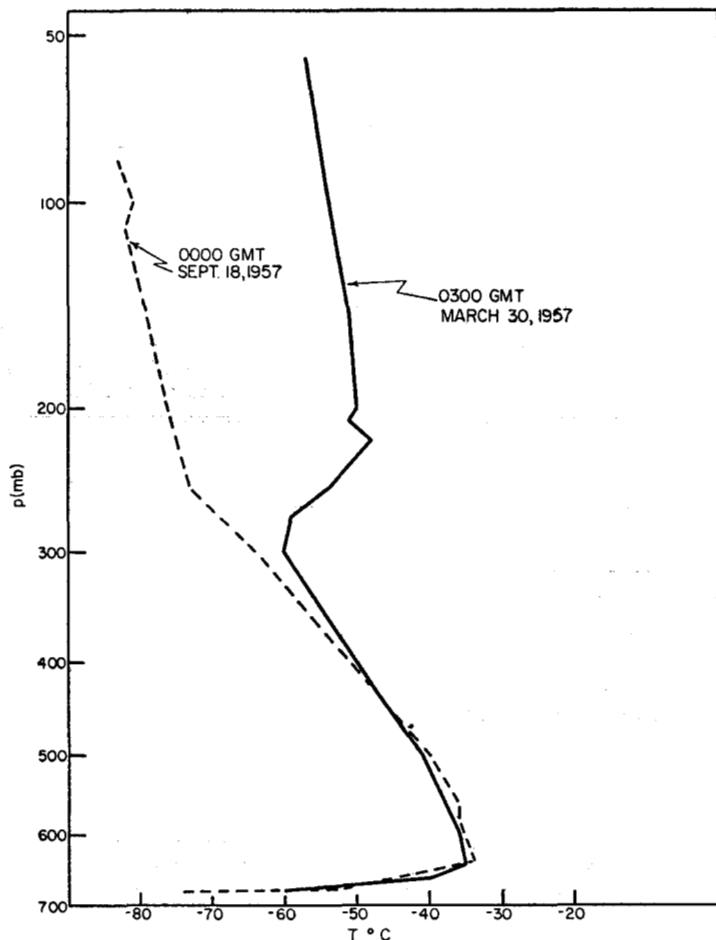


FIGURE 1.—Upper air soundings, IGY South Pole Station, at beginning and ending of the 1957 winter season.

Hence, the model is one in which, after an initial instant of time (0300 GMT, March 30, 1957; see figure 1) adjustment of the snow surface and tropospheric temperature regimes proceeds as a result of the influence of the radiation exchange processes taking place within the troposphere, between the snow surface and the inversion layer near the ground, and the radiative loss to outer space. From the initial time, and through the subsequent cooling stages, it is assumed that the troposphere is clear but saturated at all levels; that the condensed water vapor falls out of the system completely; and that the effects of the release of latent heat of condensation are negligible.

Aside from the fact that it was desired to "maximize" the tropospheric and surface cooling, the latter assumption may also be justified on the following basis. At the low temperatures which were observed in the initial sounding, the troposphere (saturated) would contain about 57.88×10^{-3} gm. of precipitable water per unit column. If during the subsequent nighttime cooling period (180 days) all of the water vapor were to be condensed out, $600 \text{ cal. gm}^{-1} \times 57.88 \times 10^{-3} \text{ gm.} = 34.73 \text{ cal.}$ would be released. This would be sufficient to warm the 670–280-mb. layer approximately

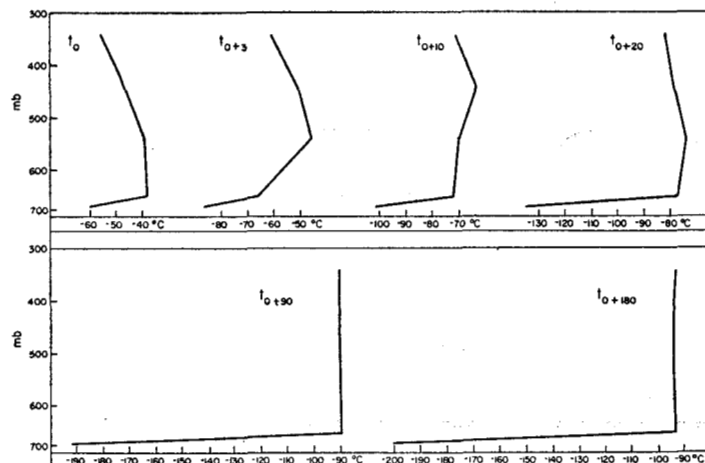


FIGURE 2.—Representative computed temperature soundings for South Pole, showing tropospheric cooling from initial time, t_0 , to 180 days later, t_0+180 .

$$\frac{34.73}{(390)(0.24)} = +0.37^\circ\text{C.}$$

It will be seen that the computed change in the mean temperature of that layer during the period is about -45°C. , or over two orders of magnitude greater.

A further simplification to reduce the number of computations involved, is to express the initial sounding as given below:

0300 GMT, March 30, 1957	
Pressure (mb.)	Temperature ($^\circ\text{C.}$)
670	-60
650	-38
540	-39
440	-47
340	-56
280	-60

As the cooling progresses, the details of the initial relative temperature distribution in the troposphere rapidly lose their identity and after the first three weeks, no vestige remains.

Since the questions to be answered cannot be expressed and solved analytically, the tropospheric and snow surface cooling was computed in a series of steps; the number of days taken in each step increased with time as the tropospheric moisture became more depleted and the cooling rates due to radiation from water vapor became smaller. Beginning from the initial time (t_0), the vertical temperature distribution in the troposphere was computed to represent the situation at 3, 10, 20, 30, 40, 50, 70, 90, 120, 150, and 180 days later. The initial and final temperature distributions as well as several intermediate ones are shown in figure 2.

3. COMPUTATIONS

The actual calculations were carried out in the following manner and approximate order:

1. Beginning at t_0 , the snow surface temperature, T_s , was assumed to fall (from -60°C.) very quickly to the temperature which gives quasi-equilibrium between the downcoming sky radiation (from water vapor and carbon dioxide) and the outgoing (black body) radiation from the snow surface, as described by Wexler [9]. For t_0 and subsequent synthetic soundings through t_{0+20} , the downcoming radiation, R_A , was determined by means of the Elsasser radiation chart. After t_{0+20} , however, R_A had to be estimated by some other means as the air temperatures exceeded the lower limits for which the standard Elsasser chart is constructed. A search was then made for a suitable empirical technique which might be used. It was found that within the range of R_A which could be computed from an Elsasser chart, a reasonably good linear relationship existed between $\log R_A$ and a temperature, T_m , defined:

$$T_m = \frac{T_0 + T_{\max}}{2} \quad (1)$$

where T_0 was the 670-mb. air temperature ($^\circ\text{C.}$) and T_{\max} the maximum tropospheric temperature ($^\circ\text{C.}$). This relationship is shown in figure 3, and the extrapolated portion (dashed line) was utilized after t_{0+20} for determining the R_A . It will be noted that the temperature defined by equation (1) is similar to the one used by Liljequist [7] in his empirical formula for R_A as a function of T_m and vapor pressure over a range of T_m from 0°C. to -45°C. Liljequist's formula was not used because it did not agree with the Elsasser chart computations in the T_m range -50°C. to -100°C. and also it gives negative values for R_A at vapor pressures less than 2.178×10^{-10} mb. (i. e., saturation temperatures lower than about -140°C.).

2. Cooling in the inversion layer, 670–650 mb., was assumed to be made up of two components: (a) the cooling effected through the net exchange of radiation with the air layers above, described in step 3, and (b) that produced by the radiative loss to the snow surface which had taken the equilibrium temperature, T_s , described above. Hence, the equation for the cooling rate due to the latter component is

$$\frac{\partial T}{\partial t} = - \frac{(\epsilon_1 + \epsilon_2) (\sigma \bar{T}^4 - \sigma T_s^4) \times 1440}{mc_p} \text{ } ^\circ\text{C. day}^{-1} \quad (2)$$

where ϵ_1 = the emissivity of water vapor in the layer

ϵ_2 = the emissivity of CO_2 in the layer

σ = the Stefan-Boltzmann constant ($.825 \times 10^{-10}$ cal. cm.^{-2} min.^{-1} deg.^{-4})

\bar{T} = the mean temperature of the layer ($^\circ\text{A.}$)

m = mass of air (~ 20.4 gm.)

c_p = specific heat of air at constant pressure (.24 cal. gm.^{-1} deg.^{-1})

Values used for ϵ_1 , for path lengths, $w \geq 5 \times 10^{-5}$ (cm. water) were those given by Brooks [1] or extrapolated from his data. For $w < 5 \times 10^{-5}$ (encountered from t_{0+90}), an

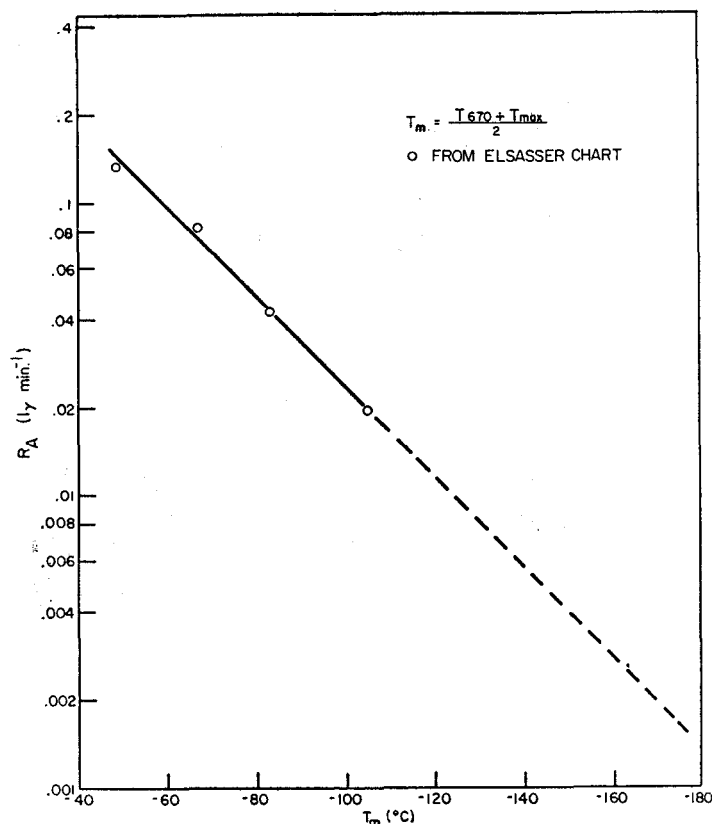


FIGURE 3.—Empirical relation between downcoming long-wave radiation and temperature, t_m .

extrapolation formula

$$\epsilon_1 = 1.69w^{1/2} \quad (3)$$

was employed. A rigorous justification for equation (3) cannot be offered, but then neither is one necessary. It has the advantage that ϵ_1 does not become zero at very small values of w , as it does in the relation given by Elsasser [2]. In the range of w where it was applied, the emissivity is so small that it hardly makes any difference for our purposes what actual value of the parameter is used. Furthermore, the radiational cooling in the layer after t_{0+20} was quite dominated by the radiation from CO_2 . For the latter a constant emissivity, ϵ_2 , of 0.09 was assumed, as given by Elsasser [2] for a 150-m. layer of atmospheric air.

TABLE 1.—Computations for t_0 sounding at South Pole (0300 GMT March 30, 1957)

p (mb.)	T ($^\circ\text{C.}$)	q (g. kg. $^{-1}$)	\bar{q}	$-\Delta p$	$\left(\frac{\bar{p}}{1000}\right)^{1/2}$	w
670	-60	.019	.120	20	.812	.0014
650	-38	.220	.230	110	.772	.0137
540	-39	.240	.180	100	.700	.0088
440	-47	.120	.090	100	.625	.0039
340	-56	.060	.053	60	.557	.0012
280	-60	.046				

Note: $w = 0.7 \bar{q} \frac{\Delta p}{\bar{p}} \left(\frac{\bar{p}}{1000}\right)^{1/2}$

TABLE 2.—Rate of temperature change due to water vapor at reference levels for sounding at t_0 .

(a) 650 mb.					(b) 540 mb.				
Level (mb.)	Δw (cm.)	$\Delta(\sigma T^4)$	$\frac{\partial \epsilon}{\partial w}$	$\Delta(\sigma T^4) \frac{\partial \epsilon}{\partial w}$	Level (mb.)	Δw (cm.)	$\Delta(\sigma T^4)$	$\frac{\partial \epsilon}{\partial w}$	$\Delta(\sigma T^4) \frac{\partial \epsilon}{\partial w}$
670-650	0-.0014	-.0820	-99.90	8.19	670-650	.0137-.0151	-.0820	-4.32	0.35
650-540	0-.0137	.0043	20.50	0.09	650-540	0-.0137	.0043	-20.50	-0.09
540-440	.0137-.0225	.0322	3.58	0.12	540-440	0-.0088	.0322	27.10	0.87
440-340	.0225-.0264	.0324	2.80	0.08	440-340	.0088-.0127	.0324	5.79	0.19
340-280	.0264-.0276	.0134	2.36	0.03	340-280	.0127-.0139	.0134	4.64	0.06
				8.51					1.38
				$\frac{\partial S_E}{\partial w}(-60^\circ, .0276 \text{ cm.}) = 0.40$					$\frac{\partial S_E}{\partial w}(-60^\circ, .0139 \text{ cm.}) = 0.83$
				$\frac{\partial T}{\partial t} = \frac{.22 \times 10^{-3} (.806) (8.91) (1440)}{.24} = -9.5^\circ \text{C. day}^{-1}$					$\frac{\partial T}{\partial t} = \frac{.24 \times 10^{-3} (.735) (2.21) (1440)}{.24} = -2.3^\circ \text{C. day}^{-1}$

(c) 440 mb.					(d) 340 mb.				
Level (mb.)	Δw (cm.)	$\Delta(\sigma T^4)$	$\frac{\partial \epsilon}{\partial w}$	$\Delta(\sigma T^4) \frac{\partial \epsilon}{\partial w}$	Level (mb.)	Δw (cm.)	$\Delta(\sigma T^4)$	$\frac{\partial \epsilon}{\partial w}$	$\Delta(\sigma T^4) \frac{\partial \epsilon}{\partial w}$
670-650	.0225-.0239	-.0820	-2.67	0.22	670-650	.0264-.0278	-.0820	-2.36	0.19
650-540	.0088-.0225	.0043	-4.19	-0.02	650-540	.0127-.0264	.0043	-3.46	-0.01
540-440	0-.0088	.0322	-27.10	-0.87	540-440	.0039-.0127	.0322	-8.26	-0.27
440-340	0-.0039	.0324	49.15	1.59	440-340	0-.0039	.0324	-49.15	-1.59
340-280	.0039-.0051	.0131	12.87	0.17	340-280	0-.0012	.0131	107.72	1.41
				1.09					-0.27
				$\frac{\partial S_E}{\partial w}(-60^\circ, .0051 \text{ cm.}) = 1.93$					$\frac{\partial S_E}{\partial w}(-60^\circ, .0012 \text{ cm.}) = 8.16$
				$\frac{\partial T}{\partial t} = \frac{.12 \times 10^{-3} (.663) (3.02) (1440)}{.24} = -1.4^\circ \text{C. day}^{-1}$					$\frac{\partial T}{\partial t} = \frac{.06 \times 10^{-3} (.583) (7.89) (1440)}{.24} = -1.7^\circ \text{C. day}^{-1}$

(e) 670-650 mb. layer					(f) Sounding for t_{0+3}				
Level (mb.)	Δw (cm.)	$\Delta(\sigma T^4)$	$\frac{\partial \epsilon}{\partial w}$	$\Delta(\sigma T^4) \frac{\partial \epsilon}{\partial w}$	Level	T (t_0)	$\frac{\partial T}{\partial t}$	T (t_{0+3})	
670-650	.0225-.0239	-.0820	-2.67	0.22	670	-60	-----	-87 [#]	
650-540	.0088-.0225	.0043	-4.19	-0.02	650	(-49)	-28	(-77)	
540-440	0-.0088	.0322	-27.10	-0.87	540	-38	-29	-67	
440-340	0-.0039	.0324	49.15	1.59	540	-39	-7	-46	
340-280	.0039-.0051	.0131	12.87	0.17	440	-47	-4	-51	
				1.09	340	-56	-5	-61	
				$\frac{\partial S_E}{\partial w}(-60^\circ, .0051 \text{ cm.}) = 1.93$	280	-60	-5*	-65	
				$\frac{\partial T}{\partial t} = \frac{.12 \times 10^{-3} (.663) (3.02) (1440)}{.24} = -1.4^\circ \text{C. day}^{-1}$					

* extrapolated from T_{450} and mean temperature for layer.

* assumed same as $\left(\frac{\partial T}{\partial t}\right)_{450}$.

3. Tropospheric cooling rates were determined by the tabular method described by Brooks [1] to give the rate of radiational temperature change at given levels of the atmosphere, in this instance the pressure levels listed in table 1 excepting 670 and 280 mb. Values for the 670-mb. level were found to be so near zero as to make it hardly worthwhile to compute them, and at 280 mb. there was too much uncertainty about the appropriate water distribution above that height. The equation used to determine w , the optical thickness of the water vapor atmosphere, was

$$w = -0.7(\bar{q}_s \Delta p / g) (\bar{p} / 1000)^{1/2} \quad (4)$$

where q_s is the saturation mixing ratio, p is pressure, and g is gravity acceleration. The "pressure-correction," $0.7 (\bar{p} / 1000)^{1/2}$, is rather a mean compromise of the factors used by various authorities; e.g., $(\bar{p} / 1000)^{1/2}$ Elsasser [2], Brooks [1]; $0.7 (\bar{p} / 1000)$, Fritz [4]; $0.4 (\bar{p} / p_0)$, Kaplan [6].

4. The cooling rates found for t_0 were applied in one step for 3 days. Then the synthetic t_{0+3} sounding was plotted, and the whole process repeated with these new data to get t_{0+10} , etc., to t_{0+180} . Selection of the number of day-units taken each time was rather arbitrary except that a temperature fall in the lowest layer of air below

the T , for the sounding was to be avoided, and some detail was desirable in order to show the transition of the sounding from its initial configuration until stabilization.

As an example, in tables 1 and 2 are given the tabulated data and computations made for t_0 ; the notation is virtually identical with that used by Brooks.

4. DISCUSSION

As shown in figure 2, the final condition arrived at was a snow surface temperature of about -200°C. , surmounted by a strong inversion in the layer adjacent to the ground, and a sensibly isothermal layer above, in agreement with the relative temperature distribution determined for much higher temperatures by Wexler [9]. The actual occurrence of such an absolute temperature distribution at the South Pole is hardly to be anticipated. However, the magnitude of the extremity of this picture would appear to be indicative of the fact that the partially closed system model upon which the calculations were based is certainly at great variance with reality. Other factors, such as the advection of warmer air into the region and the advent of cloudiness to modify the radiation loss from the snow surface, must play important roles in the development of the snow-troposphere tem-

perature regimes during the course of the long period of darkness.

Eddy conduction of heat from the inversion will also retard the cooling at the surface, although at the expense of the internal energy of the troposphere unless replaced by advection. Liljequist's [7] investigations of this flux, Q_a , at Maudheim offer an insight into the probable order of magnitude of this term. During clear strong nighttime inversions over the smooth snow fields at that location, he found the eddy flux of heat to the surface to be proportional to the wind speed at the reference level of 10 m., or,

$$Q_a = 0.0058 u_{10} \text{ (ly min.}^{-1}\text{)} \quad (5)$$

with u expressed in m. sec.⁻¹. If $u = 5$ m. sec.⁻¹, then $Q_a = 0.0290$ ly. min.⁻¹, compared to the computed radiative loss to the surface from water vapor and CO₂ in the inversion layer at t_0 of 0.0087 ly. min.⁻¹.

Some heat will also be gained by the surface during the polar night by conduction from the underlying snow. A rough estimate of the general order of magnitude of this term is also possible. From snow temperature measurements at a depth of 10 m. (where it is assumed the effects of the annual temperature cycle at the surface are negligible), Siple (communication to Wexler) places the mean annual surface temperature at about -51°C . with an annual range of the order of 67°C . If the classical heat-conduction model for a semi-infinite medium with a periodic variation of surface temperature (e. g., see Ingersoll et al., [5]) is applied to the snow layer (mean density and specific heat about 0.441 and 0.56 cgs, respectively) there would be a net gain at the surface of about 1800 cal., or an average of 0.007 ly. min.⁻¹ from the time of maximum to that of minimum surface temperatures, assumed to be 6 months apart.

In figure 1 is also shown the sounding made at 0000 GMT, September 18, 1957 near the end of the polar winter, a few hours after the new world record low temperature (at screen level) of -74.5°C . was set at the South Pole [3]. If the two soundings can be considered representa-

tive of the vertical temperature structure at the beginning and end of the polar night, it is apparent that the winter-time radiative energy losses are compensated in the troposphere by advective gains. Above about 300 mb. in the ozonosphere, however, it would appear that similar energy-balancing processes are not equally operative, as noted by Moreland [8] from analyses of data available at the Little America V Weather Central.

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